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Annual water balance and seasonality of evapotranspiration in a Bornean tropical rainforest

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Abstract

This study presents the results of 2 years combined field measurements of water vapor exchange with the atmosphere and simplified model calculations at Lambir Hills National Park, Sarawak, Malaysia (4°12′N, 114°02′E). The study site was located in a lowland mixed dipterocarp forest, a major type of Bornean tropical rainforest, where there is no clear seasonality in environmental factors such as radiation, temperature, vapor pressure deficits and precipitation; instead, unpredictable dry spells often occur throughout the year. Modified [Priestley, C.H.B., Taylor, R.J. (1972). On the assessment of surface heat flux and evaporation using large-scale parameters, Mon. Weather Rev. 100, pp. 81–92] and equilibrium evaporation expressions enabled us to understand further the environmental control of water vapor exchanges with the atmosphere and to produce a complete gap-filled data set of the hydrologic fluxes within this environment. The equilibrium evaporation from a well-watered surface sufficiently reproduced the transpiration rate (T_r); any discrepancies between the equilibrium and actual evaporation rates were caused by unpredictable intra-annual dry spells, which reduced transpiration. There were some discrepancies during the study period because of exceptional dry sequences, but in normal years the annual transpiration rate can be obtained from the equilibrium evaporation expression. The estimated annual T_r (1193.1 mm) and evapotranspiration (1545.0 mm) rates were nearly identical to the highest values reported for some humid tropical forests. Although, there were exceptional dry sequences during the study period, the annual average fraction of available energy dissipated by T_r at this site (0.69) was almost the same as the medium value between the dry and wet seasons in Amazonian tropical forests. This implies

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that the annual dynamics of the latent heat flux for this tropical rainforest are under more humid conditions than those of other tropical rainforests.

Keywords: Eddy-covariance; Evapotranspiration; Equilibrium evaporation; Priestley–Taylor equation; Tropical rainforest; Southeast Asia

1. Introduction

Tropical rainforests are among the most important biomes because of their vast amounts of primary productivity, and water and energy exchange with the atmosphere. Although these forests now cover only 12% of the global total land surface (FAO, 1993), they contain about 40% of the carbon in the terrestrial biosphere (Skole and Tucker, 1993) and are responsible for 50% of terrestrial gross primary productivity (Grace et al., 2001). Tropical rainforests are also a major source of global land surface evaporation (Choudhury et al., 1998) and have profound influences on global and regional climates and hydrological cycling (e.g. Lean and Warrilow, 1989; Nobre et al., 1991). These large latent energy fluxes from tropical rainforests are known to influence global atmospheric circulation patterns (Paegle, 1987). In the humid tropics, climate change might drastically alter hydrological regimes, for example, because of altered rainfall patterns and land cover transformation mainly as a result of forest conversion (Bruijnzeel, 1996). Consequently, such alterations might accelerate further global climate changes.

A number of field studies have reported on the components of the hydrologic cycle in Amazonian tropical rainforests (e.g. Shuttleworth et al., 1984; Shuttleworth, 1988; Hodnett et al., 1995; Grace et al., 1995; Leopoldo et al., 1995; Malhi et al., 2002; Vourlitis et al., 2002). These studies suggest that 50% of the incident precipitation in these areas is recycled as precipitation. Water cycling in Southeast Asian tropical rainforests is, however, less certain and field studies on the exchange of water between Southeast Asian tropical rainforests and the atmosphere have been poorly documented compared to the studies performed in Amazonian tropical rainforests.

As Dykes (1997) pointed out, there is a need for detailed long-term studies on the hydrological components of the ‘‘maritime’’ environments of the Southeast Asian tropics, which are quite different from the environments of the South American and Central African tropics. While there are definite periodic dry periods in the South American and Central African tropics, the Southeast Asian maritime environments do not have such phase-locked dry periods because their climate is the combined result of a summer monsoon from the Indian Ocean, a winter monsoon from the Pacific Ocean and South China Sea, and the Madden and Julian Oscillation (MJO). Northern Borneo experiences only small seasonal solar radiation and air temperature variations and annual rainfall is distributed evenly throughout the year (Inoue et al., 1993). However, unpredictable intra-annual dry spells (Kumagai et al., 2004b) and dry sequences do occur (Kumagai et al., 2001). Intra-annual dry spells, in particular, occur annually and often throughout the year. The goal of this study is to provide information regarding Southeast Asian tropical rainforest energy and water cycling. Towards this goal, measurements of hydrological components such as transpiration rate, rainfall interception and soil moisture content were initiated in 2001 as part of a project entitled ‘‘Research and Observation on the Mechanisms of Atmosphere–Ecosphere Interaction in Tropical Forest Canopy’’ in a tropical rainforest in Lambir Hills National Park, Sarawak, Malaysia (Nakashizuka et al., 2001). Transpiration rates, canopy conductances and decoupling coefficients during dry spells and in wet conditions were previously compared (Kumagai et al., 2004b). While canopy conductance and its sensitivity to micrometeorological variables were higher in wet conditions than during dry spells, these higher values did not cause higher transpiration rates compared to the dry spells. Thus, we expected transpiration in the study site to be related to the equilibrium evaporation rate at sufficiently large fetch under a steady energy supply (Slatyer and McIloy, 1961; McNaughton and Jarvis, 1983; Raupach, 2001). The present paper used the results of a further 2 years of measurements to address the following: (1) how transpiration rates are related to equilibrium evaporation in this tropical forest, (2) how the possible environmental controls of variations in transpiration respond to little climatic seasonality, and (3) how large the surface-atmosphere exchange of water vapor in a truly humid tropical rainforest is.

2. Study site and measurements

2.1. Site description

The experiment was carried out in a natural forest in Lambir Hills National Park (4°12′N, 114°02′E), 30 km south of Miri City, Sarawak, Malaysia (Fig. 1). Mean annual rainfall at Miri Airport, 20 km from the study site, for the period 1968–2001 was around 2740 mm with some inter-annual variation; for example, the maximum and minimum annual rainfall during this period were 3499 mm in 1988 and 2125 mm in 1976, respectively. Fig. 2 shows the annual rainfall cycle obtained from monthly measurements above the canopy at the study site from July 2001 to June 2002. This data was compared with the mean and standard deviations of the monthly rainfall from 1968–2001. During this time period, the driest months were February and March and the wettest period was between October and January; however, March and January are not always the driest and wettest periods, respectively, and the timing of the driest period is unpredictable. Exceptional dry sequences were observed during the course of the present study from July to August 2001 and May to June 2002 (Fig. 2).

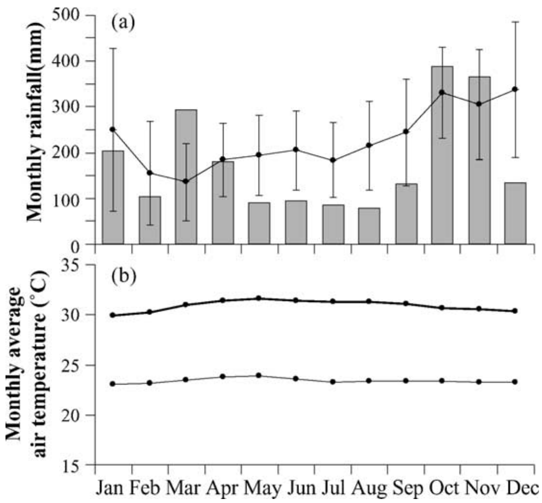

Fig. 2. (a) Monthly rainfall at the study site from July 2001 to June 2002 (bars). The points with error bars show the means ± standard deviations of the monthly rainfall at Miri Airport from 1968–2001. (b) Monthly mean maximum (thick line) and minimum (thin line) daily air temperatures at Miri Airport from 1968–2001.

A global precipitation map was reproduced using data from the Climate Precipitation Center (CPC) Merged Analysis of Precipitation (CMAP; Xie and Arkin, 1997) for the period 1979–2001. A low-pass filter was used to eliminate all periods shorter than half a year from these precipitation time series (Duchon, 1979). The ratios of the variance of the low-pass filtered precipitation to that of the unfiltered precipitation are shown in Fig. 3. The Lambir Hills National Park had one of the

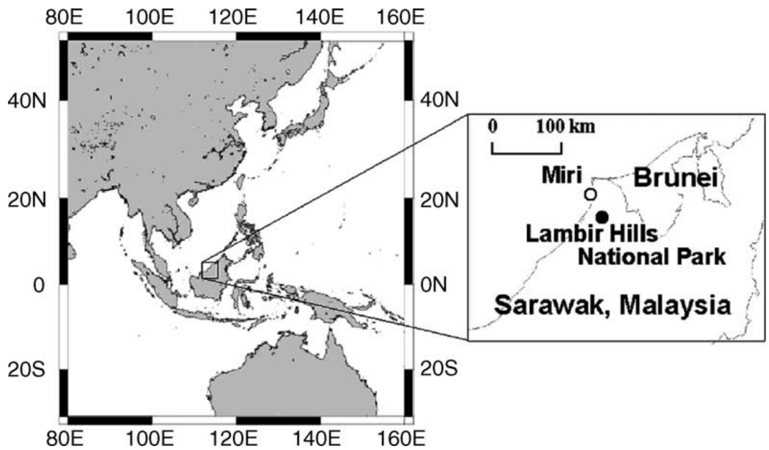

Fig. 1. Location of the Lambir Hills National Park, Sarawak, Malaysia.

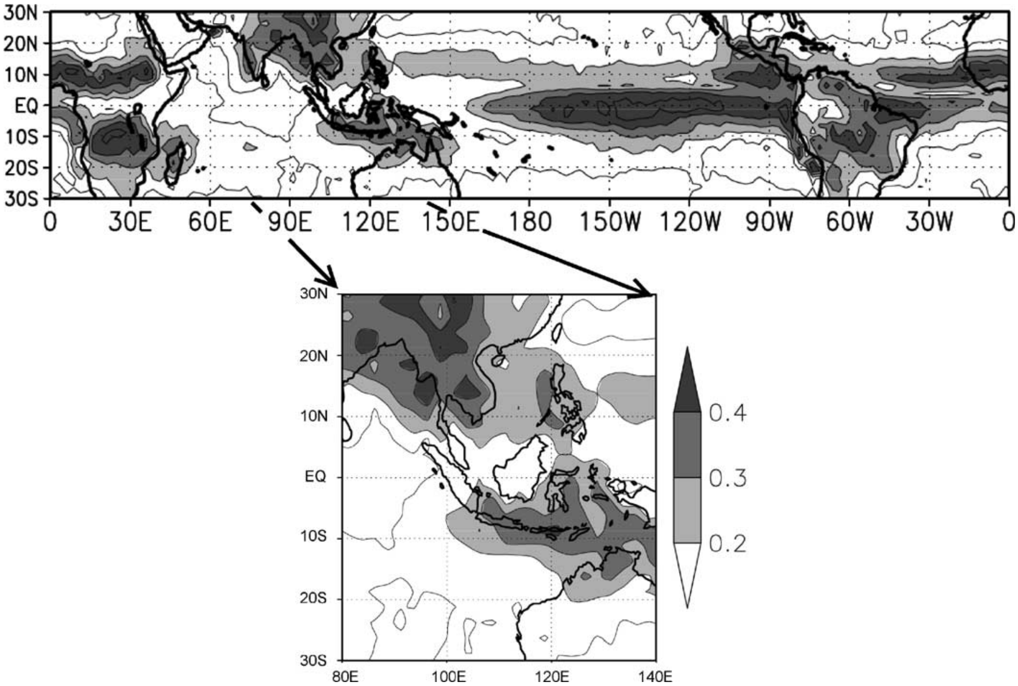

Fig. 3. Ratios of the variance of the low-pass filtered precipitation time series to that of the unfiltered precipitation. A global precipitation map was reproduced using data from the Climate Precipitation Center (CPC) Merged Analysis of Precipitation (CMAP; Xie and Arkin, 1997) for the years 1979–2001.

lowest ratios, that is, one of the weakest seasonal variations in precipitation. The mean monthly maximum and minimum daily air temperatures at Miri Airport from 1968–2001 are also shown in Fig. 2. Overall, this data suggests that the mean annual temperature is around 27 °C with little seasonal variation.

The rainforest in this park consists of two types of original vegetation common to the whole of Borneo, that is, mixed dipterocarp (Dipterocarpaceae) forest and tropical heath forest (Yamakura et al., 1995). The former contains various types of dipterocarp trees and covers 85% of the total park area. The soils consist of red–yellow podzonic soils (Malaysian classification) or ultisols (USDA Soil Taxonomy), with a high sand content (62–72%), an accumulation of nutrients in the surface horizon, a low pH (4.0–4.3), and a high porosity (54–68%) (Ishizuka et al., 1998; Sakurai, 1999). Borneo has a primarily hilly topography and most of its soils are ultisols. Ultisols are also the typical soil type throughout much of Southeast Asia, and its distribution corresponds to that of lowland and hill dipterocarp forests (Ohta and Effendi, 1992).

2.2. The crane facility

A 4 ha experimental plot at an altitude of 200 m, gridded into 400 subplots or quadrats of $10 \times 10\ m^2$, was used in this project. An 80 m tall (at the base of the gondola) canopy crane with a 75 m long rotating jib was constructed in the center of this plot to provide access to the upper canopy. Observational stages at four levels (23.5, 38.5, 59.0, and 75.5 m above the ground) accessible by an elevator were also installed. The 59.0 m high stage was devoted to Eddy-covariance flux measurements. The subplots or quadrats were used for in-canopy micrometeorological measurements, and throughfall and stemflow measurements. The canopy height surrounding the crane is about 40 m but the height of the emergent treetops can reach up to 50 m.

2.3. Leaf area measurements

The leaf area index (LAI) was measured in 2002 using a pair of plant canopy analyzers

(LAI-2000, Li-Cor, Lincoln, NE) at 5 m intervals in a $30 \times 30\,m^2$ subplot. The LAI ranged from 4.8 to $6.8\,m^2\,m^{-2}$ with a mean of $6.2\,m^2\,m^{-2}$ (Kumagai et al., 2004b). The monthly amounts of litterfall were evenly distributed throughout the year (M. Nakagawa, unpublished data) perhaps suggesting small variations in LAI.

2.4. Micrometeorological and soil moisture measurements

The following instruments were installed at the top of the crane, 85.8 m from the forest floor: a solar radiometer (MS401, EKO, Tokyo, Japan), an infrared radiometer (Model PIR, Epply, Newport, RI), and a tipping bucket rain gauge (RS102, Ogasawara Keiki, Tokyo, Japan). Upward short-wave radiation and long-wave radiation were measured at a separate tower located 100 m south of the crane using an upside-down solar radiometer (CM06E, Kipp & Zonen, Delft, Netherlands) and infrared radiometer (Model PIR, Epply, Newport, RI), which have been recording since 15 December 2001. Radiation was sampled every 5 s and an averaging period of 10 min was used (CR10X datalogger, Campbell Scientific, Logan, UT). These measurements were used to compute the net radiation (R_n; Wm^{-2}); the net radiation time series prior to 15 December 2001 were calculated using the relationship between solar radiation (R_s; Wm^{-2}) and net radiation ($R_n = 0.87R_s-35.0$, $R^2 = 0.99$).

The volumetric soil moisture content (θ) and matric potential (ψ) were measured in a subplot at 10, 20 and 50 cm below the forest floor at 10 min intervals (CR23X datalogger, Campbell Scientific, Logan, UT). A time-domain reflectometer (TDR; CS615, Campbell Scientific, Logan, UT) was used to measure the time series of θ and a tensiometer (DIK3150, Daiki, Tokyo, Japan) was used to monitor ψ at the same depth as θ thereby providing the necessary measurements to compute in situ soil water retention curves at each depth. The relative extractable water in the soil (Θ; $m^3\,m^{-3}$) was calculated as follows using the average θ in the 0–50 cm soil layer: $(\theta - \theta_r)/(\theta_s - \theta_r)$, where θ_s and θ_r are the saturated and residual water contents, respectively. θ_s was taken as the maximum average θ observed in the 0–50 cm soil layer during the experimental period (June 15 2001–November 10 2002), and θ_r was estimated as the average θ when ψ at a depth of 10 cm was bounded by -0.7 m and -0.01 m. The Θ results range from 0 (soil dry) to 1 (soil saturated).

2.5. Evapotranspiration measurements by the Eddy-covariance method

A three-dimensional sonic anemometer (DA-600 (Kaijo, Tokyo, Japan) from 15 June 2001 to 4 August 2002, and USA-1 (METEK, Elmshorn, Germany) since 4 August 2002) was installed at 60.2 m, that is, 20 m above the mean canopy height. An open-path CO_2/H_2O analyzer (LI-7500, Li-Cor, Lincoln, NE) was placed about 30 cm away from the sonic anemometer in an inclined orientation and with its windows coated in wax to reduce problems associated with rain and dew droplets. In addition, a ventilated psychrometer (MS020S, EKO, Tokyo, Japan) was installed adjacent to the sonic anemometer and open-path analyzer to calibrate and correct the density effects (Webb et al., 1980).

The sensors were mounted 2.5 m away from the crane leg using a horizontal boom fixed on the southwest corner, so as to minimize flow distortion when the wind was blowing from its prevailing northwesterly and southeasterly directions. When the crane was constructed, a gap to hold the crane at its center was formed. The largest radius of the gap was about 25 m, and the fetch exceeded 1 km in the northwest sector; the minimum fetch, which was in the southwest sector, was about 200 m. The influences of the gap and minimum fetch on the flux estimations were examined using a two-dimensional analytic solution of the diffusion equation proposed by Schuepp et al. (1990) together with the stability dependency proposed by Lloyd (1995). As a result, we found that the wind almost always flowed from the northwest during the daytime, and that the maximum 80% effective fetch and the minimum 10% effective fetch were about 1500 m (12:30 h, 18 June 2001) and 50 m (2:00 h, 13 June 2001), respectively. Therefore, these influences were neglected for flux estimations during the daytime.

Wind speeds and gas concentrations were sampled at 10 Hz and the averages were computed at 30 min intervals. The results were detrended using a moving-average time constant of 200 s and the wind field (u, v, w) (ms^{-1}) coordinates were rotated so that the mean v

and w values were zero over 10 min periods. The percentage corrections for the coordinate rotations were less than 1%, probably because of the gentle slope. We applied the density fluctuation correction method described by Webb et al. (1980). In this study, the flux data measured during rain events were abandoned because data from the Eddy-covariance system obtained during rainfall events are not reliable. However, as the durations of the rain events during the study period were short, gaps occurring in the flux data because of this abandonment were small. Calibrations of the open path analyzer were performed periodically using a dew point generator (LI-610, Li-Cor, Lincoln, NE): the zero and span were checked at intervals of 2–3 months.

2.6. Throughfall and stemflow measurements

A plot representative of the 4 ha experimental plot was instrumented for throughfall and stemflow measurements. Throughfall was measured using 40 collectors composed of manually-operated gauges assembled with 20.6 cm diameter plastic funnels and 10 L capacity bottles, 20 of which were dispersed on a grid over the fixed plot. Samples were taken manually, everyday except Sunday between 8:00 and 13:00 LST using 1 L plastic cylinders with 10 ml graduation intervals.

For stemflow measurements, collars connected to 30 L bottles or 15.7 cc tipping bucket flow meters (no. 34, Ohta Keiki, Tokyo, Japan) were attached to the trunks of 81 trees with various circumferences (ranging between 1 and 130 cm). Of these trees, 78 were located inside the fixed plot and three outside. Some stemflow measurements were sampled as the time of tip by the event data logger (KADEC-UP, Kona System, Sapporo, Japan); all others were measured manually using the same procedure as for the throughfall measurements. Details on the throughfall and stemflow measurements are presented in Manfroi et al. (2004)

3. Methods for estimating evapotranspiration

Equilibrium evaporation occurs when air passes over an extensive wet surface and becomes saturated. Then, the Bowen ratio ($\beta = H/LE$ where H and LE are the sensible and latent heat flux, respectively) takes the value γ/Δ where Δ is the changing rate of saturation water vapor pressure with temperature (PaK^{-1}), and γ is the psychrometric constant (66.5 PaK^{-1}). Under this condition, the daily equilibrium evaporation (LE_{eq}, MJm^{-2} d^{-1}) was obtained using the energy balance equation as follows:

$$LE_{eq} = 0.0864\frac{\Delta}{\Delta+\gamma}R_n \qquad (1)$$

where R_n is the daily net radiation above the canopy (Wm^{-2}). Here, for the sake of simplicity we assumed that the daily cumulative storage heat flux to the soil and canopy could be neglected. The thermodynamic variable Δ was calculated based on the mean air temperature averaged over daylight hours. Daily net radiation was obtained by averaging these hourly values.

Although the precipitation, solar radiation, and air temperature at the study site show small seasonal variations, intra-annual dry spells that often occur throughout the year have been reported (Kumagai et al., 2004b). Therefore, it is logical to expect that environmental changes caused by this unpredictable dry condition, especially, changes in soil moisture, will affect the biological activities in this site; here the resultant changes in transpiration are examined. To determine transpiration we need to link its limiting factors to the variation in soil moisture in the rooting zone. Hence, we used a modified Priestley and Taylor (PT) (1972) expression to compute the daily transpiration rate (LE, MJm^{-2} d^{-1}) according to Scanlon and Albertson (2003) as follows:

$$LE = \eta\alpha_p LE_{eq} \qquad (2)$$

where α_p is the PT coefficient (1.26) and η is the limiting factor. De Bruin (1983a) found that α_p varies with surface resistance and the entrainment rate of dry air at the top of the atmospheric boundary layer (ABL). By assuming that surface resistance is related to soil moisture, we used the Eddy-covariance LE measurement to obtain the daily η value then proceeded to derive a relationship between Θ and η. Here, the averaged η was computed in bins of Θ (incremented in 0.1) and a regression model describing Θ and the bin-averaged η was derived ($R^2 = 0.96$) as follows:

$$\eta = 0.46\Theta + 0.53, \quad \text{for } 0 \leq \Theta \leq 1.0 \qquad (3)$$

The values of the α_p and η products obtained here ranged from 0.6 to 0.8 and appeared reasonable when compared to other dry canopy studies (De Bruin, 1983a).

In Kumagai et al. (2004a), daily interception (I_c, mm d^{-1}) was related to daily rainfall P (mm d^{-1}) using the following equation:

$$I_c = \begin{cases} P, & \text{for} \quad 0 \leq P < 0.90 \\ 0.13P + 0.78, & \text{for} \quad 0.90 \leq P < 3.91 \\ 0.084P + 0.96, & \text{for} \quad 3.91 \leq P \end{cases} \qquad (4)$$

To estimate I_c, we assumed that a single storm event does not exceed one day, and that multiple storm events during 1 day can be lumped together as a single daily storm. This assumption is necessary because interception losses are related to daily precipitation thereby forcing a daily minimum time step of integration. We believed this simplifying assumption was reasonable here because single storm events exceeding 1 day seldom occur, at least in the historical records of Miri Airport.

4. Results and discussion

4.1. Energy balance

Fig. 4 compares the sum of the daily cumulative sensible (H) and latent (LE) heat fluxes ($H + LE$) to the daily cumulative net radiation (R_n) for dry and wet canopy conditions. The mean differences between the $H + LE$ and R_n values were 2.72 and 1.81 MJm^{-2} day^{-1} on days with no rainfall events and with both no rainfall and rainfall events, respectively. The student's paired t-test indicated that the differences between these quantities for both conditions were significantly different from zero at a 0.01 probability level (t_s = 10.67 and $t_{s0.01}$ = 2.63 on days with no rainfall events, and t_s = 9.82 and $t_{s0.01}$ = 2.60 on whole observation days). The $H + LE$ value was regressed against the R_n value, and the average slope was calculated as 0.78 ± 0.03 (95% confidence) on days with no rainfall events and 0.82 ± 0.04 on whole observation days. The $H + LE$ values on days when rainfall occurred tended to be larger than those on days when no rain events occurred, despite discarding the Eddy-fluxes measured during rainfall events. This might have been caused by evaporation from the wet canopy, which were included in water vapor flux measurements obtained on days when rainfall occurred.

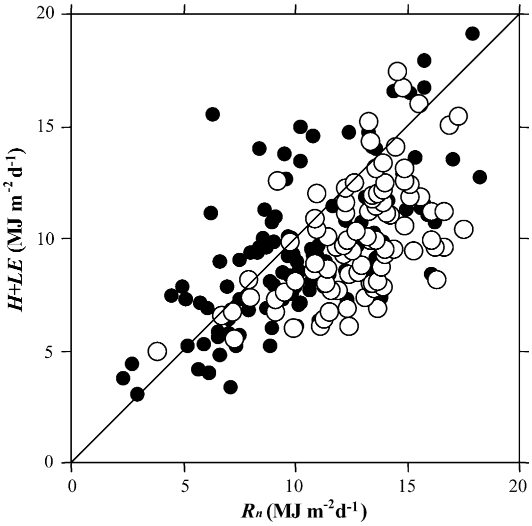

Fig. 4. The sum of the daily cumulative sensible (H) and latent (LE) heat fluxes ($H + LE$) vs. the daily cumulative net radiation (R_n) for a dry (open circle) and wet canopy (closed circle).

On a one-dimensional scale, the energy balance was not completely closed here; a possible source of error in the energy balance might be attributed to the different areas that were sampled for the $H + LE$ and R_n measurements. On the other hand, non-closure of the energy balance is a common feature of Eddy-covariance measurements above forest canopies, and the closure deficit observed here (about 20%) was comparable to that obtained at most forested sites (e.g. Goulden et al., 1996; Ohta et al., 2004; Aubinet et al., 2001; Wilson et al., 2002). This imbalance could be explained by a variety of processes such as an inadequate performance of the Eddy-covariance system or scale mismatches between the Eddy-fluxes and micrometeorological samples (e.g. McMillen, 1988). The energy balance relationship at this site most likely indicates that the Eddy-fluxes were underestimated by about 20%; an underestimation that was caused in part by insufficient evaluation of the contribution of the relatively large-scale Eddy (Lee, 1998; Malhi et al., 2002). Although further analyses are needed to resolve the reasons for this imbalance and the Eddy-flux underestimations, the degree of closure was well within the range reported for other forested sites. Therefore in this study, according to Kumagai et al. (2004b), we assumed that the Eddy-fluxes had been equally underestimated and thus multiplied them by 1/0.78.

4.2. Estimating evapotranspiration

Fig. 5 compares the LE_{eq} (see Eq. (1)) calculated using independent measurements of R_n and air temperature against the measured LE for dry canopy conditions. Note that the LE_{eq} and LE from a dry canopy are synonymous with transpiration (T_r). A significant (t_s = 3.09; $t_{s0.05}$ = 1.99) but small (0.9 MJm^{-2} d^{-1} or 10% of the mean) difference existed between the LE_{eq} and measured LE. Overall, LE_{eq} clearly reproduced the measured LE value despite its simplified form.

The first definition of LE_{eq} was evaporation from a wet surface into saturated air (Priestley, 1959). However, air above extensive wet surfaces such as the ocean is seldom saturated (Priestley and Taylor, 1972). McNaughton (1976) presented a more general idea of LE_{eq} suggesting that it occurs whenever the ABL has a constant height and is capped by a lid through which no exchange of air occurs. Furthermore, McNaughton and Jarvis (1983) showed that LE_{eq} is evaporation in the limit of complete decoupling.

A previous paper by the authors reported high decoupling coefficient values under wet soil moisture conditions due to relatively large surface conductance compared with aerodynamic conductance (Kumagai et al., 2004b). This might explain why in this study site the LE_{eq} can reproduce the measured LE value. Fig. 6 compares the LE_{eq} and LE values for dry canopy conditions modeled using the modified PT expression (see Eq. (2)). The LE_{eq} values were clearly correlated with the modeled values (R^2 = 0.74), but Fig. 6 also reveals a systematic overestimate caused by the limiting factor responding to dry soil moisture conditions in the modified PT expression. However, despite the existence of dry conditions and the overestimates, the mean difference between the LE_{eq} and modeled values was 0.8 MJm^{-2} d^{-1}, or 8% of the population mean.

Many researchers have suggested that LE_{eq} is a guide (even sometimes as a good as an approximation) of actual evaporation rates over well-watered vegetated surfaces, even though ideal field conditions seldom occur (Raupach, 2001). While there were some dry conditions in the observation period that could have affected LE (see Fig. 2), the measured LE somewhat conformed with the LE_{eq}.

LE values measured on rainy days were excluded from the T_r modeling. However, we assumed that

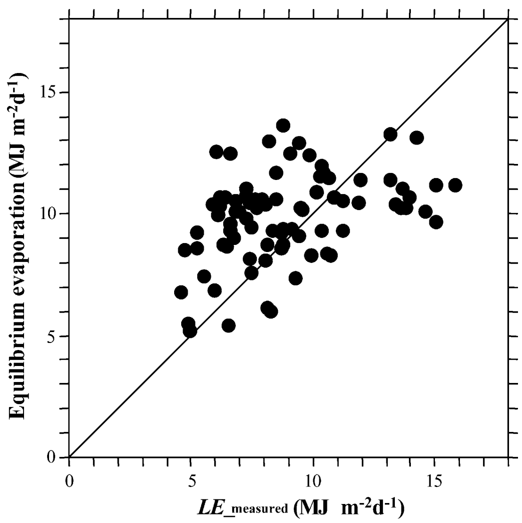

Fig. 5. Comparison between the measured latent heat flux ($LE_{\_measured}$) and equilibrium evaporation.

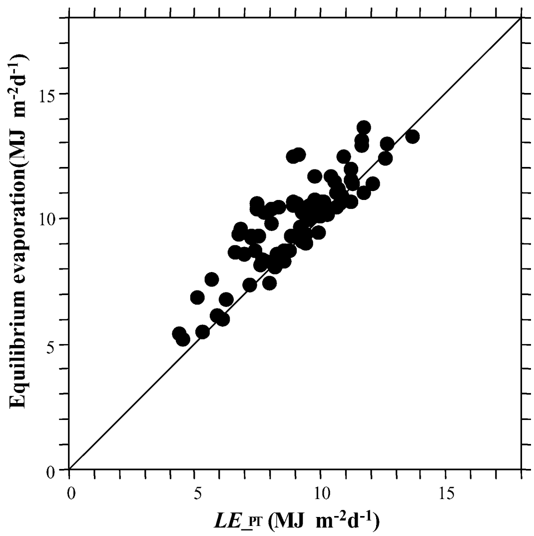

Fig. 6. Comparison between the latent heat flux estimated using the modified Priestley and Taylor expression ($LE_{\_PT}$) and equilibrium evaporation.

modeled T_r rates could also be calculated on days when rainfall occurs, because the duration of storm events are short in this site and the modeled values were calculated from the daily cumulative R_n. Thus, the LE on rainy days was calculated as the sum of the T_r modeled by Eqs. (1) or (2) and the I_c modeled by Eq. (4). This calculation is analogous to the model used by Shuttleworth and Calder (1979) for actual evaporation that includes both the transpiration and evaporation rates of intercepted water; we filled data gaps and replaced unreliable data using this calculation.

4.3. Annual water balance

The annual variation in water balance is summarized in Table 1, which shows the monthly totals and averages. The monthly total transpiration rates (T_{r_eq} and T_{r_pt}) were estimated from the complete gap-filled data set obtained using the equilibrium evaporation (Eq. (1)) and modified PT (Eq. (2)) expressions, respectively.

During months with little rainfall, namely, August 2001, September 2001 and June 2002, the T_{r_eq} were about 20 mm larger than the T_{r_pt}. This implies that the T_{r_eq} expression does not take into consideration the effect of dry soil moisture conditions. As a result the annual total value of T_{r_eq} was 94 mm larger than that of T_{r_pt}. However, the equilibrium evaporation expression for estimating T_r is still effective because the annual total rainfall in the study period (July 2001–June 2002) was exceptionally low (see Fig. 2). Furthermore, for practical estimations it has the advantage of simplicity because it requires only net radiation and air temperature as inputs.

Having verified that Eq. (3) describes the effect of soil moisture conditions on T_r reasonably well and that the T_{r_pt} expression consequently reproduces the measured T_r value, we proceeded to determine the seasonal and annual water cycles using T_{r_pt}. T_r was controlled by the "trade-off" between precipitation (P) and average daily net radiation (R_n); even if there is plenty of rainfall, T_r is small when R_n is small, and even if there are many dry days, if there is plenty of R_n, T_r can be large.

Here, the monthly daily average latent heat flux (LE) was estimated from the monthly total evapotranspiration rates, that is, T_r plus the monthly total interception (I_c). Therefore, since I_c effectively increased with P, the R_n fraction dissipated by LE (fr) surpassed 100% in months with much rainfall (October and November 2001). fr ranged between 0.65 and 1.22, suggesting a close correlation between P and fr (Table 3). While most fr results reported in other tropical forests did not take into consideration evaporation from the wet canopy, our results included the effect of the wet canopy in energy partitioning. When we calculated fr using only T_r, it ranged between 0.57 and 0.78 and the annual average value was 0.69. In this study, the annual average fr estimated from T_r was nearly identical to the value in the transition period between the wet and dry seasons in a transitional tropical forest (0.65; Vourlitis et al., 2002)

Table 1
Monthly summary of the total precipitation (P), transpiration (T_r), interception (I_c), the average daily net radiation (R_n), and fraction of R_n dissipated by latent heat flux (fr)

	P (mm)	T_{r_eq}/T_{r_pt} (mm)	I_c (mm)	R_n (MJ/m^2 d)	$fr_{\_eq}/fr_{\_pt}$
Jul 2001	84.5	118.9/110.9	16.3	13.24	0.70/0.65
Aug 2001	79.0	115.7/98.9	15.4	11.88	0.87/0.76
Sep 2001	131.0	114.7/96.6	26.6	13.38	0.86/0.75
Oct 2001	389.0	96.9/100.0	55.0	10.04	1.19/1.22
Nov 2001	365.0	90.0/89.9	51.1	9.87	1.16/1.16
Dec 2001	135.0	100.1/99.4	29.3	10.84	0.94/0.93
Jan 2002	204.5	102.6/99.2	27.6	11.34	0.90/0.88
Feb 2002	103.5	84.2/71.0	23.6	10.34	0.91/0.80
Mar 2002	294.0	112.8/102.4	37.5	11.70	1.01/0.94
Apr 2002	179.5	125.9/125.6	26.7	13.36	0.93/0.93
May 2002	91.0	118.0/110.7	23.3	12.01	0.92/0.88
Jun 2002	94.5	107.2/88.5	19.5	11.42	0.90/0.77
Year Total	2150.5	1287.1/1193.1	351.9	11.62	0.94/0.89

The subscripts eq and pt designate the values estimated using Eqs. (1) and (2), respectively.

and the mean value of the wet and dry seasons in an Amazonian tropical rainforest (0.73 and 0.54 in the wet and dry seasons, respectively; Malhi et al., 2002). It is notable that although there are no clear seasonal variations in rainfall in this site normally, because of the exceptional dry months in the study period our lowest estimated *fr* value was comparable to that found in Amazonian tropical forests during the dry season (0.54: Malhi et al., 2002; 0.5: Vourlitis et al., 2002). Our results imply that the annual dynamics of the latent heat flux for this tropical rainforest are under more humid conditions than for the above tropical rainforests.

The annual T_r and I_c in 2001–2002 were estimated as 1193.1 and 351.9 mm, respectively, suggesting that I_c can be up to 30% of T_r (Table 1). The ratio of I_c to total P was 16%. Evapotranspiration, computed as the sum of T_r and I_c, was 1545.0 mm, which accounts for about 72% of the total P. Evapotranspiration studies by Bruijnzeel (1990) in humid tropical forests suggest that: (1) the average annual T_r was 1045 mm (range 885–1285 mm), (2) the average I_c was 13% of the incident P (range 4.5–22%), and (3) the annual evapotranspiration ranged from 1310 to 1500 mm. While our interception losses were comparable to this reported value, our T_r and evapotranspiration values were on the highest boundary of their range for tropical forests (e.g. Calder et al., 1986; Bruijnzeel, 1990; Leopoldo et al., 1995). According to De Bruin (1983b), our annual evapotranspiration value can be seen in the evapotranspiration category from stations with a long wet season and it can therefore be determined with the PT method.

5. Conclusions

In this Bornean tropical rainforest, the equilibrium evaporation from a well-watered surface sufficiently reproduced the transpiration rate. The discrepancies between the equilibrium and actual evaporation rates were caused by unpredictable intra-annual dry spells, which reduced transpiration. There were some discrepancies in this study period because of exceptional dry sequences, but in normal years the annual transpiration rate can be obtained from the equilibrium evaporation expression. This also suggests that further research is needed with regards to the effect of inter-annual variability in the length of dry periods, especially at this site because it is a maritime environment where the effects of El Niño-Southern Oscillation (ENSO) and monsoon strength are prominent.

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